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Earthquakes and tsunamis caused by low-angle normal faulting in the Banda Sea, Indonesia

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As the world's largest archipelagic country in Earth's most active tectonic region, Indonesia faces a significant earthquake and tsunami threat. Understanding this threat is a challenge because of the complex tectonic environment, the paucity of observed data, and the limited historical record. Here we combine information from recent studies of the geology of Indonesia's Banda Sea with GPS observations of crustal motion and an analysis of historical large earthquakes and tsunamis there. We show that past destructive earthquakes were not caused by the supposed megathrust of the Banda Outer Arc as previously thought, but are due to a vast submarine normal fault system recently discovered along the Banda Inner Arc. Instead of being generated by coseismic seafloor displacement, we find the tsunamis were

more likely caused by earthquake-triggered submarine slumping along the fault's massive scarp, the Weber Deep. This would make the Banda Detachment representative not only as a modern analogue for terranes hyper-extended by slab rollback, but also for the generation of earthquakes and tsunamis by a submarine extensional fault system. Our findings suggest that low-angle normal faults in the Banda Sea generate large earthquakes, which in turn can generate tsunamis due to earthquake-triggered slumping.

Subduction zones generate the world's largest earthquakes and the vast majority of large, destructive tsunamis. The Banda Sea is underlain by one of the world's most striking subduction zones, with a concave-westward arc bending 180° in a tight 300 km radius of curvature (Figure 1). It would at first appear likely that large earthquakes and tsunamis that devastated the Banda Islands in the historical past[1,2], as well as the potential threat of future events[3–5], should be attributed to a megathrust along this Banda Outer Arc. However, since the Banda Arc is a zone of arc–continent collision, it no longer features an oceanic trench – and therefore, no megathrust[6–9]. Hence, it is imperative that the mechanism for destructive Banda Sea earthquake and tsunami generation is re-evaluated.

The configuration of the Banda subduction zone has long been contested by proponents of models involving either the bending of a single slab[10], or subduction of two separate slabs from opposing directions[11]. More recently, it has been proposed that the evolution of the arc is best explained by rollback of a single slab into a pre-existing embayment in the Australian continental margin that controlled development of the slab's tight curvature[9]. A major implication of this scenario is that the active tectonics of the Banda Sea would be dominated not by thrusting but by extension, as lower crust and subcontinental lithospheric mantle is exhumed to fill the gap opened above the rolling-back slab. Recent field evidence from Seram and the eastern Banda arc strongly supports this hypothesis[12–15]. This extreme rollback-driven lithospheric extension has also been shown to account for the most intriguing physiographic feature of the Banda Sea: The Weber Deep[16].

At 7.2 km depth, the Weber Deep is the deepest point of the Earth's oceans not within a trench. The eastern wall and floor of the Weber Deep have been recognized as the scarp of a vast but previously undocumented low-angle normal fault (LANF) system, the "Banda Detachment", which has been the primary structure facilitating upper-plate extension in the Banda Sea[16]. Since the forearc extension that formed the Weber Deep commenced at c. 2 Ma[18] and the Weber Deep is 120 km wide, the implied average geologic slip rate of the Banda Detachment is about 6 cm/yr. Furthermore, it is notable that little sediment has accumulated in the Weber Deep, less than 1 km thickness based on seismic reflection data[19,20]. These two observations imply the Weber Deep, and therefore the Banda Detachment, must be young features, even though no focal mechanisms determinable from any seismic catalogue are consistent with earthquakes on this low-angle fault[16]. This may be because the detachment slips aseismically, or during frequent low-magnitude events, or during rare, large-magnitude earthquakes, the most recent of which must have occurred prior to the modern seismic record.

Normal fault slip on the Banda Detachment should be detectable using Global Positioning

System (GPS) measurements of crustal motion, but eastern Indonesia's complex tectonic setting and paucity of observations make this challenging. While GPS measurements of vertical motion can be directly used to assess fault-related motion, in order to identify such effects in observations of horizontal motion they must be referenced to a local tectonic framework. We developed a tectonic block model for the Banda Sea area (Methods and Figure S1) and estimated motion of the Banda Block without using data from the Banda Islands, where an uplift rate of 1.4 mm/yr (Table ST1) suggests fault-related strain accumulation. Then we subtracted the inferred block motion from the observations in the Banda Islands. These residual horizontal motions, as well as the observed uplift rate at Band Neira, are consistent with interseismic locking of a normal fault aligned with the Banda Detachment (Figures 2 and S1). In this sense, the available GPS measurements of crustal motions suggest interseismic strain accumulation associated with normal faulting along the Banda Detachment. However, the limited data available are likely consistent with other interpretations.

Below we consider whether historical accounts of destructive earthquakes in the Banda Sea can be explained by large but infrequent earthquakes on the Banda Detachment, and how these might generate tsunamis. We focus on the earthquake of 26 November 1852, because it has the most extensive and detailed accounts of the shaking that devastated the Banda Islands and of the subsequent tsunami[21,22].

1 What is the source of the 1852 Banda Sea earthquake?

While the Banda Sea is an area of high seismicity[9,23], the vast majority of large, instrumentally recorded earthquakes are $>$50 km depth intraslab events (Figure 3a), that are weakly if at all felt in the Banda Islands. The world's largest intraslab event ever recorded, the 1938 Mw=8.5 Banda Sea earthquake, was felt only weakly in the Banda Islands and did not cause a large tsunami[24]. No earthquake since the 19th century has caused significant damage or deaths in the Banda Islands[25].

By contrast, historical accounts from the 17-19$^{\text{th}}$ centuries document at least 5 earthquakes that caused widespread destruction in the Banda Islands: 1683 - "most houses became rubble heaps"; 1710 -" most houses were damaged irreparably"; 1763 - " Three-quarters of all houses of Banda Neira were transformed to rubble heaps"[21,22]. These and other earthquakes felt strongly in the Banda Islands were often accompanied by ground cracking/fissuring, tsunamis, and prolonged sequences of felt aftershocks, none of which are typical of intraslab earthquakes. These are all characteristics of shallow earthquakes, so we conclude that, unlike the large earthquakes in the Banda Sea recorded since the beginning of the 20$^{\text{th}}$ century, the historical earthquakes that caused damage and fatalities in the Banda Islands were shallow.

Accounts of the 1852 Banda Sea earthquake and tsunami are particularly detailed[21,22,26]. A summary of the accounts and our interpretation of them in terms of Modified Mercalli Intensity (MMI) are given below and shown in Table ST2 and Figure 3. Figure 3a shows that the earthquake generated its strongest felt intensity in the Banda Islands, which we have assigned MMI 8, and intensity decreases northward to MMI 4 at Ternate. Felt reports from eastern Java previously ascribed to this event[1,21] appeared to imply an enormous felt area, but these observations are more likely associated with a Mw 5.7-6.0 earthquake on the Pasuruan Fault in eastern Java[27]. Similarly, the emergence of a small island in the Kai Archipelago observed in 1853 was thought to indicate coseismic displacement in the rupture area of the 1852 earthquake[1,21], but we interpret this observation to be a mud volcano eruption, which are prevalent in the Kai Islands[28] and can be triggered by earthquakes even at great distance[29]. We therefore discount the Kai Islands mud volcano as indicative of the rupture area of the 1852 earthquake.

The felt area we consider for the 1852 Banda earthquake is therefore more restricted than that of previous studies[1] (Figure 3b). We used a grid search for the source parameters of the 1852 earthquake, applying Bayesian inference to characterise uncertainties[30]. The results are shown in Figure 3c-d, indicating that the high-probability zone for earthquake locations that best explain the intensity data extends to the north and east of the Banda islands, with magnitudes in the range 7.5-8.7. The only major fault identified so near the Banda Islands is the Banda Detachment, and we therefore consider whether an earthquake on this fault, just east of the Banda Islands (Figure 3a, red rectangle) could give rise to the observed seismic intensities. We have considered an earthquake at the lower end of the confidence interval in magnitude, Mw 7.5, located along the surface trace of the Banda Detachment, since this is more likely to have a fault dip that could rupture in an earthquake - i.e., 12° near the scarp[16], with the steep bathymetry increasing the effective dip to 18°.

In Figure 3b we compare the ground motion calculated for this Mw 7.5 earthquake scenario to that for a Mw 8.4 megathrust rupture of the Tanimbar Trough. For both earthquakes we use the same subduction interface Intensity Prediction Equation (IPE)[31], because it is based on MMI observed in a subduction zone setting and accommodates large earthquakes on shallowly-dipping faults (to our knowledge there is no IPE for large LANF earthquakes). Figure 3b shows that even a very large earthquake on the Tanimbar Trough or elsewhere on the Banda "megathrust" is too far away to produce intensities as strong as those observed: the 1852 earthquake must have been not only large, but very close to the Banda Islands. In order to produce the rapid fall-off in intensities northward, towards Ambon, Seram and Ternate, the rupture area must have been relatively compact; a much larger rupture area in the Tanimbar Trough[1] generates intensities that do not decrease sufficiently with distance northward. Instead, the observed intensities favor a smaller earthquake near the Banda Detachment.

2 What is the source the 1852 Banda Sea tsunami?

Any tsunami in the Banda Islands generated by an earthquake on the supposed outer arc megathrust, whether the Seram Trough to the north or the Tanimbar Trough to the south, would have negative polarity (i.e., "draw-down"). This is a consequence of the arc-inwards dip of the fault, which generates a pattern of vertical seafloor displacement that is downwards in the direction of the Banda Sea and upwards along the rim of the outer arc. This can be seen from the tsunami waveforms calculated for megathrust earthquakes that were thought to have caused the 1629[2] and 1852[1] tsunamis, which have pronounced draw-downs as first-arriving tsunami energy (Figures 8 and 10 of the respective papers, and our Figure 4c). The four tsunami observations of the 1852 event all clearly indicate positive polarity, followed by rapid draw-down of sea level, showing that the source could not have been a megathrust event in the outer arc.

The account of the 1852 tsunami in Banda Neira includes a particularly clear description of its arrival time relative to the earthquake: after "vertical shocks ... of 5 min duration", " the ground had been calm for a quarter of an hour when a flood wave crashed in"[21,26]. This 20 min delay time between the occurrence of the earthquake and the arrival of the tsunami is an important constraint on the locus of tsunami generation. In Figure 1 we show an inverse tsunami travel time map, which shows where a tsunami would have originated had it arrived at Banda Neira at various times following an earthquake . The 20 min contour of this map highlights two potential locations where the tsunami could have originated: (1) The Banda Detachment, where it emerges on the western side of the Weber Deep about 100 km SSE of Banda Neira; and (2) a large submarine slump on the the Weber Deep's eastern side (WDS in Fig. 1b).

We modelled tsunami generation by coseismic seafloor displacement due to a large number of scenario earthquakes rupturing the Banda Detachment at the potential source location SSE of Banda Neira. Although the extremely low dip ($\approx 8^{\circ}$) of the Banda Detachment results in a tsunami with clear positive polarity, we found that even very large earthquakes (Mw 8.4) produced tsunami heights that were smaller than those observed (Figure 4c - red curves). The rupture area of such a large earthquake would necessarily extend into the deeper part of the Banda Detachment where the dip is essentially zero[16], and we therefore regard this tsunami generation mechanism as unlikely.

When considering a submarine slump on the eastern scarp of the Weber Deep, we were guided by the extensive slump scarp, of about 100 km along-scarp length and 50 km down-scarp width (WDS in Figure 1b). It is the largest of at least 4 such scarps evident on both west and east sides of the Weber Deep and its deepest edge coincides with the 20 min inverse travel time contour, so its triggering at the time of the earthquake should match the observed arrival time. We simulated slump-generated tsunamis using a two-layer approach[32,33], finding that the scenario that best matches the observations is a slump 40 km long by 15 km wide (Figure 4a, WDS-11), and of 50 m thickness (i.e. volume 30 km^3), which results in the tsunami waveforms at Banda Neira shown in Figure 4c (blue curves).

The slump-generated tsunami waveforms in Figure 4c (blue curves) have positive initial polarity followed by a rapid draw-down, which matches the historical accounts. At Saparua, the tsunami height builds over several cycles to 3 m, while at Banda Neira the second peak is highest at 5.5 m, giving a peak-to-peak sea level variation of 7.5 meters that matches the observations well (the sailing vessel "Hai", anchored in 11 m water depth before the tsunami, saw this depth decrease to 7 m on arrival of the tsunami, then later increase to 14.5 m[22]; it was also reported that "the difference between the highest and the lowest water lever was 26 feet [8.2 m]"[21]). The reported sea level variations at Ambon are more ambiguous, but not inconsistent with the simulated 1.5 m height. The tsunami waveforms simulated for the slump source match the observations much better than those of the coseismic displacement source (Fig. 4c), and we therefore conclude that a slump was most likely the cause of the 1852 Banda Islands tsunami.

3 Other historical Banda Sea earthquakes and tsunamis

Can the mechanism for earthquake and tsunami generation of the 1852 event apply also to other historical earthquakes in the Banda Sea? As discussed above, in the 17-19$^{\text{th}}$ centuries, at least 4 other earthquakes have caused widespread destruction in the Banda Islands. These earthquakes did not generate felt reports from elsewhere, were often accompanied by ground cracking/fissuring and prolonged sequences of felt aftershocks, and in some cases caused tsunamis. All of these factors argue for a shallow source of major earthquakes near the Banda Islands, and the Banda Detachment is the only known active fault large enough to support such earthquakes. For this reason, we suggest that the Banda Detachment is likely to be the source of not only the 1852 earthquake but also the four other earthquakes known to have devastated the Banda Islands.

It is more speculative to suggest that other major tsunamis that have affected the Banda Islands, in 1629, 1763 and 1841, were caused by earthquake-triggered submarine slumps. However, the propensity for accumulations of sediment along the edges of the Weber Deep to slump down its steep slopes is evidenced by several large slump scars on both western and eastern sides of the Weber Deep (Figure 1). The one identified here as a potential source of the 1852 tsunami is the largest, but there are at least three others, two on the western and one on the eastern side. The other tsunamis associated with the historical Banda Islands earthquakes could be associated with these slumps, or it could also be that the slump we have suggested as the source of the 1852 tsunami occurred in multiple stages. For the tsunami of 1763 it is reported that: " During the first shocks, the Sea level fell 9 m (30 feet) and then quickly rose (in less than 3 minutes)"[22]. This initial draw-down of sea level could be associated with a slump scar on the western side of the Weber Deep, whose polarity would be opposite that of tsunamis generated by slumps on the eastern side, and which is much closer to Banda Neira than the WDS (see Figure 1).

4 LANF rupture, slump triggering and the earthquake catalog

Activity and seismicity on LANFs has been controversial, since there are few examples of LANF earthquakes in the seismic record, and the mechanics of LANF slip are difficult to explain[34–36]. While occurrence of a Mw 7.5 event on a LANF would be the largest ever considered, we note that large LANF earthquakes are not without precedent: Earthquakes as large as Mw 6.8 have been documented in New Guinea's Woodlark Basin[37,38] and Mw 6.4 in the western Gulf of Corinth[39]. The Banda Detachment is by far the largest-known and potentially most active LANF in the world, and is the only known fault near the Banda Islands large enough to host earthquakes capable of causing extensive damage. On the other hand, the earthquakes we associate with the Banda Detachment do not necessarily have to have occurred on the low-angle detachment itself. It is possible that they were confined or at least nucleated on a more steeply-dipping normal fault above the Banda Detachment that has yet to be identified[40,41].

To understand the propensity for submarine slumps in the Weber Deep to generate tsunamis, Figure S2 displays several cross sections across the Weber Deep, in which maximum slopes are calculated on either side of the basin. Maximal slopes range from 3-14°, with half being greater than 6°. A study of earthquake triggered submarine slumps along the eastern continental slope of the USA[42], found that earthquakes of Mw 7.5 can trigger submarine slumps at greater than 150 km distance - about the distance from our hypothesized Banda Detachment earthquake to the slump scar on the eastern slope of the Weber Deep – for slopes greater than 6°, a threshold distance which increased rapidly for steeper slopes. While this result depends on properties of the sediment and depth to the slump failure plane, it suggests that the possibility of earthquakes on the Banda Detachment triggering slumps on the steep sides of the Weber Deep is not unrealistic.

Finally we address the question of why, if the Banda Detachment is a major source of earthquake and tsunami hazard, is there no evidence of earthquakes rupturing the Banda Detachment in available earthquake catalogs? The same question could have been raised regarding lack of seismicity on the Sumatra megathrust prior to the occurrence of the 2004 Sumatra-Andaman earthquake. In the case of Sumatra, a series of earthquakes in the mid 19th century was followed by a period of quiescence throughout the 20th century, until the Sumatra megathrust "re-awakened" in 2004. It could be that the same is true of the Banda Detachment, that the series of destructive earthquakes and tsunamis from the period 1629-1852 was followed by a long period of seismic quiescence. The same has been noted for Java, where despite the occurrence of many large, destructive earthquakes in 1681-1877, only one has occurred since[30]. Regardless of which fault caused the Banda Sea earthquakes of 1629-1852, it would be a mistake to assume the Banda Detachment can't rupture in a future earthquake simply because it lacks recorded seismicity.

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Methods

Tectonic block modelling of GPS velocities. We have combined the published GPS velocities[43–46] for eastern Indonesia with recent additional GPS measurements at stations in the Banda Sea region (Table ST1). We modelled these as a sum of rotations and elastic strain [47] along boundaries of the tectonic blocks depicted in Figure S1, using information from published studies[43,48,16,12]. The rapid subsidence and minimal sediment accumulation in the Weber Deep[16] suggests active, distributed deformation there that could drive normal slip on the Banda Detachment but is not accounted for by block modelling. We therefore exclude from the block modelling stations BAPI and BANI, whose proximity to the Banda Detachment may cause them to be affected by this strain accumulation. The block model's estimate of BAND's pole of rotation with respect to AUST is at longitude 123.16°, latitude 0.02° with a clockwise rotation of 3.4°/Myr, resulting in SSW motion of BAND with respect to AUST, and in left-lateral slip along the Kawa Fault cutting through the island of Seram that is consistent with its current geomorphological expression[12]. This left-lateral sense of slip continues along the Banda Detachment, with a slight normal component along its northern section (BAND-SERA) and a thrust component on the southern (BAND-TIMOR). While the movement of BAND does not itself result in normal relative plate movement along the Banda Detachment, if we subtract the calculated from the observed velocities at BAPI ad BANI, we obtain residual velocities at both sites directed NW (see inset Figure S1), in an up-dip direction along the same Banda Detachment rupture we used to model the reported ground motions in the 1852 earthquake. The direction and magnitude of these horizontal velocities at BANI and BAPI, as well as the uplift rate at BAPI (no uplift estimate is available for BAPI), are consistent with interseismic locking of the Banda Detachment, as indicated in Figure 2.

The residual horizontal velocities at BANI and BAPI, and the uplift at BANI, are aligned with calculated velocities due to interseismic locking of a shallow-dipping normal fault (Figure 2 and S1) along the Banda Detachment with a slip rate of 5 cm/yr, slightly less than the 6 cm/yr implied by the opening of the Weber Deep to 120 km since 2 Ma[18]. While this implied fault slip is normal in sense, we note that this direction is not consistent with the ESE direction of opening of the Weber Deep, and stress that more data is needed to uniquely resolve the sense of motion on the Banda Detachment. The low subsidence rates observed in the outer arc at CAUL and CSAU are consistent with weaker strain accumulation in the fold-and-thrust belt.

Seismic intensity (MMI) inversion The intensity data for historical earthquakes in the Banda Sea are too few to constrain relationships between magnitude, distance, and intensity[49]. Instead, source parameters for the 1852 earthquake are estimated by undertaking a grid search of source parameters using ground-motion models (GMMs) and ground motion to intensity conversion equations (GMICEs) to forward model intensity, and Bayesian inference is applied to calculate probability of source parameters given the historical intensity data[30]. Initially, we calculate the root-mean-squareerror (rmse) between the modeled and observed intensities for each parameter combination[49]. Since there are large uncertainties inherent in the poor sampling, assignment of intensity values, and use of GMMs and GMICEs, *a priori* information is given as uniform over all parameters except magnitude for which a Gutenberg–Richter prior distribution is used with a b-value of 1. This choice reflects our knowledge that the greater rate of small earthquakes means it is more likely a given observed intensity is due to a small earthquake generating anomalously strong shaking than to a large earthquake generating anomalously weak motion. As yet there are no Indonesia-specific GMMs or GMICE, so we used a combination of GMMs[50–53] (with weights 0.5, 0.25, 0.125, and 0.125 respectively) and GMICE[54] derived for tectonically active environments elsewhere.

Tsunami modelling. Tsunami modelling is conducted using the JAGURS tsunami simulation code[55]. The code numerically solves non-linear shallow water wave equations in a spherical coordinate system using a staggered-grid, finite-difference scheme. The tsunami simulations are performed on a domain with nested grids, with the coarsest and finest grid resolutions approximately 450 and 50 m, respectively (Figure S3). A time-step of 0.2 s is set to satisfy the Courant stability condition, and the digital elevation model (DEM) used was built from combining the Indonesian National Bathymetry (BATNAS)[56], a marine chart around the Banda Islands[57], and SRTM-90m[58].

The slump on the eastern side of the Weber Deep is simulated using a 2-D, two-layer flow model in which the upper layer corresponds to the ocean and the lower layer to a turbidity current, with the time-varying seafloor displacement so obtained used as a boundary condition for the tsunami simulation[33]. To simulate the slump as a turbidity current, both layers use the long wave approximation, with flow velocities integrated in the vertical direction, hyrdostatic pressure of the ocean layer applied to the top of the lower layer, and an interfacial shear stress applied between the layers proportional to the square of their differential velocity[32]. The slump layer is initiated as a Gaussian-shaped function superposed on the current bathymetry, elongated along the top of the lanslide scarp visible on the eastern wall of the Weber Deep (WDS in Figure 1b). We considered a wide variety of lengths (10-75 km), aspect ratios (2-10) and thicknesses (50-300 m). The scenario that best matched the observations was 40 km long by 15 km wide (Figure 4a, WDS-11), with a thickness of 50 m (i.e. volume 30 km^3).

Data availability

All of the seismic intensity observations used here are based on historical accounts available in the published literature[21,22,26]. Except for the newly estimated velocities in Table ST1, all of the GPS velocities are available from published sources[43–46]. The raw GPS data on which the new velocities in Table ST1 are based can be obtained from the Indonesian Geospatial Information Agency (BIG). The elevation data used for tsunami modelling is a combination of the Indonesian National Bathymetry (BATNAS)[56] (see http://tides.big.go.id/DEMNAS/, last accessed in June 2019), a marine chart around the Banda Islands[57], and SRTM-90m[58].

Code availability

All of the codes used in this study have been described in published work and are available in the public domain. Tectonic block modelling was accomplished using the software TDEFNODE[47], availabe at http://www.web.pdx.edu/ mccaf/defnode.html (last accessed August 2019). The EQIAT code[30] used for Bayesian inference of earthquake parameters from seismic intensity data is available at https://github.com/GeoscienceAustralia/EQIAT (last accessed August 2019). Earthquake ground motion modelling was performed using Openquake[59], available at https://github.com/gem/openquake (last accessed August 2019). Finally, the tsunami modelling used the JAGURS[55] software available at https://github.com/jagurs-admin/jagurs.

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Author Contributions

P. R. Cummins has led the writing of the paper, undertaken the ground motion modelling and supervised the analysis of historical accounts, the tsunami modelling, and tectonic block modelling of the GPS data.

I. R. Pranantyo conducted the tsunami modelling and analysis of historical accounts.

J. M. Pownall provided the analysis of geologic evidence for slab rollback and of the evidence for slumping in the bathymetry data.

J. D. Griffin undertook the Bayesian analysis of the historical intensity observations.

I. Meilano analysed the raw GPS position data to determine crustal velocities.

S. Zhao conducted the tectonic block motion analysis of the GPS observations.

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Figure 1 The tectonics of slab rollback in the Banda Sea region. (a) Tectonic setting of the Banda Sea, indicating the relative positions of the Tanimbar and Seram Troughs in the outer arc, the Banda Detachment along the inner arc, and the Weber Deep. Contours are of inverse tsunami travel-time from Banda Neira, KSZ is the Kawa Shear Zone, and WDS is the location of the submarine slump scarp modelled as a potential source of the 1852 tsunami. White dashed lines indicate other slump scarps as well as striations in the Weber Deep along the direction of slab roll-back, and X-X' indicates the locations of the cross sections in (c). (b) An inset showing in greater detail the bathymetric signature of the slump scar (WDS) used to guide the modelling of the 1852 Banda Islands tsunami. (c) The Banda Slab and the Banda Detachment profile along X-X' in (a) (modified from [16]).

Figure 2 Crustal movement and faults in the Banda inner and outer arcs. (a) GPS observations of crustal motion along an idealized profile through the Banda Inner and Outer Arcs. Residual horizontal velocities at BAPI and BANI and vertical velocities at BANI in the Banda Islands are compared with modelled velocities[17] for interseismic locking of a shallow-dip (12°) normal fault slipping at 5 cm/yr. Vertical motions at CUAL and CSAU are also indicated as possibly due to locking of faults in the Tanimbar-Seram fold-and-thrust Belt. Figure S1 shows the locations of BAPI, BANI, CUAL and CSAU (note that vertical motions are not available for BAPI, and residual horizontal motions for CUAL and CSAU are not shown because their horizontal velocities were used in the block modelling). (b) A conceptual profile normal to the Banda Detachment, from inside the Banda Inner Arc to just outside the Banda Outer Arc, illustrating the relationship between the the Banda Detachment and Tanimbar-Seram fold-and-thrust belt used to model the GPS velocities in (a) with the hyper-extended lithosphere underlying the Weber Deep. The black symbols denoting fault slip indicate the sense of long-term-fault slip; note that these are opposite in sense to the “backslip” applied to model the interseismic deformation.

Figure 3 Banda Sea earthquakes and seismic intensity modelling. (a) Earthquake activity in the Banda Sea, including hypothesised Banda “megathrust” rupture areas for the 1629 (dashed: modelled by [2]; solid: modified to follow the actual deformation front), and 1852[1] earthquakes. Red box is the Banda Detachment rupture area for the 1852 earthquake proposed here. (b) Modelling of observed intensities vs rupture distance using a subduction interface IPE[31] for the Mw 7.5 Banda Detachment (orange) and Mw 8.4 Tanimbar Trough[1] (blue) models for the 1852 Banda Sea earthquake. Note that triangles represent the intensities assigned to the historical accounts by [1], whereas circles are the generally more conservative assignments made in this study. (c) and (d), posterior distributions for location and magnitude, respectively, of the 1852 Banda Sea earthquake, based on Bayesian inference for observed intensities. Large stars indicate the most probable parameters from the posterior distribution, small stars indicate the least-squares solution, and dashed lines in (d) indicate the 95% confidence interval for magnitude.

Figure 4 Selected tsunami models of the 1852 Banda Sea event; a) A Mw=8.4 earthquake on the Banda Detachment (BD Mw8.4), a Mw=8.4 earthquake on the Tanimbar

Trough[1] (TT Mw8.4), and a slump on the eastern side of the Weber Deep (WDS-11) – note that the scale bars are different; b) Simulated maximum tsunami height from scenarios in a; c) Simulated tsunami waveform at three virtual gauges where detailed historical accounts are available. The red curves in (a) are contours of the tsunami inverse travel times.